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SOLAR ENERGY AND THE SEASONAL THERMOCLINE

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SUMMARY

The decrease in underwater light intensity with depth is primarily a function of absorption by the water of the sun's light. This phenomenon can be studied by underwater photometry, which makes it possible to calculate the amounts of energy absorbed per unit time and per unit volume at a given depth. In this way one can derive the annual cycle of the "theoretical" vertical temperature profiles corresponding to the absorption of solar energy. This solar heating gives rise to a density gradient resulting in the vertical stability of the water.

/1.4-1*

Heat exchanges with the atmosphere cause a cooling of the surface layer, which results in a local loss of stability which can be propagated to deeper waters by turbulent thermal conduction. Wind contributes to this process by a mixing to varying depths. There is a progressive obliteration, starting from the surface, of the initial theoretical profile, and its replacement by a homothermal layer whose temperature and thickness vary with time. Beneath this mixed layer the seasonal thermocline appears, at the same depth at which the theoretical profile (and the stability of the water) reappears.

* Numbers in the margin refer to pagination in the foreign text.

Thus the temperature profile results from solar heating and surface cooling due to heat exchanges with the atmosphere. As these phenomena are linked, it is possible to determine the thickness of the mixed layer from a knowledge of the solar energy flux and thermal exchanges with the atmosphere (or the resulting surface temperature). It is always possible for advection by ocean currents to upset the situation.

INTRODUCTION

The decrease in temperature with depth is not uniform and for almost the entire year there appears in the vertical profile a layer of maximal thermal gradient called the "thermocline". This thermocline, located relatively close to the surface, is called seasonal because its annual cycle results from changing solar energy input. This temperature gradient also appears as a sharp vertical density gradient which acts as a barrier to momentum and heat propagation, as well as to the diffusion of particles and of substances in solution. Furthermore, this temperature gradient causes substantial modification in the transmission of underwater sound waves. Various authors have proposed schemes of varying degrees of complexity to explain this phenomenon.

A considerable simplification results if one assumes that turbulent thermal conduction becomes negligible at the level of the thermocline. Then the temperature profile of the first hundred meters results directly from the influx of solar energy at various depths and surface exchanges with the atmosphere, this latter effect being limited to the mixed layer above the thermocline. We will give, by way of example, one of the results of a heat balance study [1] based on hydrologic and meteorological data from the Liguro-Provencal basin, in the north-east of the western

Mediterranean (Fig. 1).

REVIEW OF THE HEAT BALANCE EQUATION

One can write the air-ocean heat exchange balance in the form

$$Q_s(1-A) = Q_t + Q_e + Q_c + Q_n + Q_{ad}.$$

$Q_s(1-A)$ is the input of solar energy reduced by a fraction corresponding to the ocean's albedo. This is the solar radiation absorbed by the water during the time interval under consideration. In our study we took the values of Q_s measured at the Meteorological Station of the Nice Airport. The curve in Fig. 2a represents the mean annual variation of Q_s for the years 1965 to 1971. As such records are in fact rather scarce, one can also make use of tables or weather atlases which give solar energy flux values in clear air as a function of latitude [2,3,4]. Using empirical formulas one can take cloudiness into account, which, in the case of Nice, involves a reduction of about 20%. The albedo of the ocean's surface (the fraction of sunlight reflected and backscattered to the atmosphere) has been studied, in particular, by a group from our laboratory [5]. It varies between 6% in summer and 8% in winter. A mean value of 7% has been proposed in another paper by M. I. Budyko in 1956.

Q_t is the variation during the same time interval of the heat content of the water. This term is calculated from the vertical temperature profiles. Its value results from the absorption of the sun's radiant energy, heat exchange across the surface, and thermal contributions from advection.

Q_e , Q_c , and Q_n are the exchanges with the atmosphere:

Q_e by evaporation, Q_c by thermal convection, and Q_n by infrared radiation (also called "nocturnal" radiation). These exchanges, occurring entirely at the surface, represent in general a loss of energy by the ocean and manifest as a cooling of the water.

Q_{ad} is the contribution of heat from marine advection. We shall return later to the effects of this term, which we shall assume for now to equal zero.

PENETRATION OF SOLAR RADIATION INTO THE OCEAN AND THE CORRESPONDING ELEVATION OF WATER TEMPERATURE

The reduction of undersea illumination with depth corresponds to the absorbed flux. If we neglect the ascending light, which only represents a few percent of the descending light, we have the familiar formula /1.4-2

$$\frac{dF}{dsdz} = -\frac{dE_d}{dz} = k E_d$$

where E_d is the descending light and k the extinction coefficient defined between the depths z_1 and $z_2 > z_1$ by the relation

$$k_{z_1}^{z_2} = \frac{\log E_{z_1} - \log E_{z_2}}{dz}$$

where E is the illumination in W/cm^2 , z is in meters, and k in m^{-1} .

In heat balance studies over varying intervals the relevant quantity is the solar energy absorbed (per unit of surface area). This quantity corresponds to the integral of illumination over time, called "irradiation" and expressed in J/cm^2 . We shall designate this quantity by the letter q . A study [6] conducted with a laboratory-buoy ($\phi = 42^\circ 2'N$, $G = 5^\circ 6'E$) in 1969 showed that the

extinction coefficient k can be considered a constant over the course of one solar day (independent of the height of the sun). However this coefficient is very variable with depth in the first few meters as a result of the selective absorption of the water, principally of the short and long wavelengths. At depths greater than 5 to 10 m it becomes essentially constant. Besides this it varies with location and with the seasons under the influence of local and transient phenomena that affect the transparency of the water. Since there are no continuous recordings of underwater illumination for each station, we have adopted an average value for extinction deduced from different measurements made in the region under study. This has yielded, for the entire year, an overall absorption of 70% in the first 5 m, and an extinction coefficient k equal on the average to $6 \cdot 10^{-2} \text{ m}^{-1}$ beyond this depth (this corresponds to the characteristics of ocean water belonging to type I or II in the classification of N. G. Jerlov [7]).

At the surface we have irradiation $Q_s(1-A) = q_0$, while between 0 and 5 m of depth we assume an overall absorption of 70%, and so the irradiation at 5 m is $q_5 = 0.3 q_0$. At any depth beyond this we have $q_z = 0.3 q_0 e^{-k(z-5)}$, and the energy absorbed between depths z and $z + dz$ is

$$q_z - q_{z+dz} = 0.3 q_0 e^{-k(z-5)} (1 - e^{-k dz}).$$

Figure 2b shows the fractions of solar energy absorbed in this manner per cm^2 of surface and per m of depth at various levels. It is not necessary to continue such a graph to depths greater than 70 or 100 meters, since the residual energy at these depths is less than a thousandth of that incident on the surface. This absorption translates into a temperature rise which is calculated taking into account the density of the water ρ , its specific heat c_p ,

and the mechanical equivalent of the calorie J (for ocean water the product $J\rho c_p = 4Jgr^{-1}\circ C^{-1}$). The curves of Figure 2b can be converted either into an increase in temperature at a given depth over the course of the year or into a variation of the vertical temperature profile at a given instant. In the latter case, choosing one initial thermal state, one can trace the changing successive profiles resulting from the absorption of solar radiation for the different time intervals considered. It is convenient to take as an initial thermal state one corresponding to the end of March. At this time in the north-west Mediterranean one finds an isothermal water layer, with a temperature very close to $13^\circ C$, extending from the surface to depths of 200 meters and beyond. Thus in Figure 3 the first curve, designated 04, represents the vertical temperature profile resulting from the solar contribution for the month of April. The succeeding curves show the cumulative effects of the following months, through December. It can be seen that the change from one profile to the next is greatest in June (maximum solar energy contribution) and least in December (minimum solar contribution). These theoretical profiles would hold if the sea were a closed system, solidified and with zero thermal conductivity.

MODIFICATIONS TO THE PROFILE DERIVED FROM SOLAR HEATING

These modifications are due mainly to the effects of exchanges with the atmosphere and mixing by the wind. Solar heating gives rise to a density gradient which imparts vertical stability to the water. Exchanges with the atmosphere cause a cooling which upsets this stability in the region of the surface and so can be propagated to greater depths by turbulent thermal conduction. The wind reinforces this process by a mixing to varying depths. There is a

progressive obliteration, starting at the surface, of the theoretical profile, which is replaced by a homothermal zone called the "mixed layer" whose thickness and temperature vary with time according to the heat balance. Under this mixed layer there is the so-called seasonal thermocline, which is a steep temperature gradient, and it is from this depth on that the solar heating profile and the stability of the water reappear. Thus, one can delineate a vertical profile based on the equilibrium between solar heating and surface cooling due to exchanges with the atmosphere. From the surface down to a certain depth (that of the thermocline) this profile represents an isothermal layer, while at greater depths it follows the theoretical profile for solar heating.

We shall now compare the computed profiles with those /1.4-3 actually found during different seasons. The temperatures shown in Figures 4a, 4b, and 4c were measured at three stations in the Liguro-Provencal basin (Fig. 1). In 4a and 4b the measurements were made along a line between Nice and Calvi (90 miles), at 15 miles from Nice in 1970 and 55 miles from Nice in 1971. The temperatures in 4c were measured in 1964-65 at the laboratory buoy about 55 miles from both Nice and Calvi. These graphs also show the profiles calculated from equilibrium considerations, using the mean solar energy input and the surface temperature of the water. Despite the various approximations (mean solar input, mean extinction coefficient) there is satisfactory agreement throughout the year between the temperatures measured and the calculated profiles in Figures 4a and 4b. In Figure 4c, showing the data from the laboratory buoy, there is a qualitative agreement for the two periods in June, when the thermocline was near the surface. In September, however, the experimental

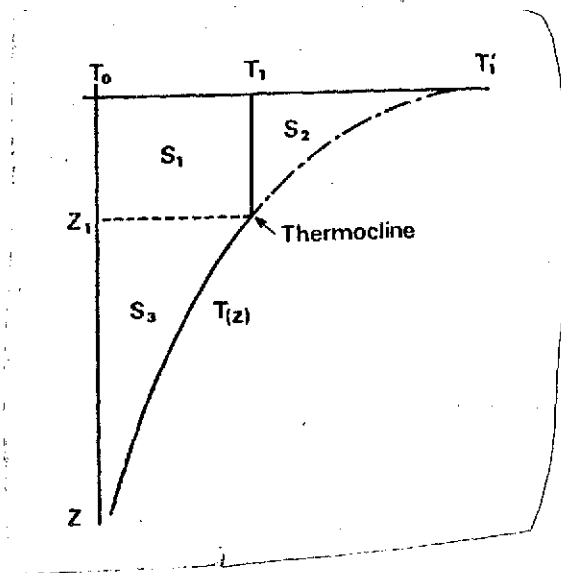
results deviate markedly from the equilibrium profile. At the other stations along the line from Nice to Calvi the observed temperatures deviated in varying degrees from the calculated profile. In our heat balance study [1], for lack of adequate meteorological coverage, we assumed the same solar energy input and the same exchange with the atmosphere over the entire Liguro-Provencal basin, which could explain part of the discrepancy. Advection effects are probably also a contributing factor.

GRAPHIC DETERMINATION OF THE DEPTH OF THE THERMOCLINE

The diagram in Figure 3, as we have explained, is derived from a knowledge of the solar input (measured directly or from an atlas of values) and of the mean extinction coefficient of the water. The ocean surface temperature value, measured directly or with infrared techniques, permits a graphic determination of the approximate depth of the thermocline throughout the year. One need only find the depth along the theoretical profile at which the temperature is the same as that measured at the surface.

CALCULATION OF THE DEPTH OF THE THERMOCLINE

In the above the surface temperature took into account the effects of exchanges with the atmosphere. Certain atlases or publications give local values for just these exchanges. It is then possible, with a few approximations, to calculate fairly simply the depth of the thermocline from this information. Thus, a method proposed by H. Lacombe [8] only requires a knowledge of the solar input and the sum of the exchanges with the atmosphere. This author assumes, as we do, that there are no advection effects; on the other hand he uses an extinction coefficient value k which is the same for all depths (even the first several few meters).



In the graph shown here the isothermal profile T_0 could correspond to that found in the sea at the end of March. The profile $T(z)$ reaches a temperature at the surface of T'_1 , which represents the theoretical profile of solar heating assuming a value of k , constant, up to the surface. The equilibrium profile, starting at the surface with a temperature T_1 and coinciding with the theoretical profile at depths $z > z_1$ results at the end of April from the solar input I_0 and losses due to surface exchanges with the atmosphere P_0 . Considering the areas of the graph S_1 , S_2 , and S_3 (which are proportional to energies) we arrive at the relation

$$(S_1 + S_2 + S_3) - (S_3) = (S_1) + (S_2)$$

$S_1 + S_2 + S_3$ corresponds to the absorption of solar radiation I_0 . S_3 corresponds to the absorption of the energy that has reached a depth of z_1 , i.e., $Iz_1 = I_0 e^{-kz_1}$. S_1 is the solar energy absorbed at depth z_1 , multiplied by the depth z_1 of the thermocline, i.e., $z_1 k I_{z_1}$. S_2 represents the cooling due to exchanges with the atmosphere, i.e., P_0 .

Thus we have the relation $I_0(1 - e^{-kz_1}) = P_0 + z_1 k I_0 e^{-kz_1}$, and defining

$$m = \frac{I_0 - P_0}{I_0}$$

and $kz_1 = q$ we can write

$$e^q = \frac{1 + q}{m}.$$

The value of m taken from an atlas then permits calculation of the depth z_1 of the thermocline (or that of the mixed layer) if one knows the extinction coefficient of the water. Because of the way in which it was derived this formula is only applicable when the depth of the thermocline is diminishing, that is, when one begins with an isothermal layer of thickness z_0 and arrives at a new layer of thickness $z_1 \ll z_0$. This condition requires that the value of m in the above formula be increasing. The maximum value of m corresponds to the time of minimum depth of the thermocline (approximately the month of June). After this time we observe an increase in the depth of the thermocline. One can no longer calculate the depth of the thermocline at a given moment, though it is possible to find the relative variation (for more details see the original article by H. Lacombe [8]).

The problem with this method is that it assumes an extinction coefficient constant for all depths. One would have to substitute a function $k(z)$ that varies in the first few meters, which would complicate the calculation. Beyond this, although the values of I_0 found in atlases can be considered reasonably trustworthy, the same is not the case for P_0 (evaporation, convection and nocturnal radiation heat loss), which varies greatly at a given point as a function of meteorological conditions. H. Lacombe has calculated the monthly temperature distribution at the laboratory buoy of which we spoke earlier; it is interesting to compare these estimates of thermocline depth with the temperature profiles observed at this location, and shown in Figure 4c. For the end of June this author would predict a depth of about 23 meters, while in reality the

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thermocline is located at about 10 meters. This discrepancy is perhaps linked to the assumption made about the extinction coefficient k . The comparison is more favorable in September, when the values are respectively 28 and 23 meters, and in December, when they are 48 and about 45 meters.

CONCLUSION

The hypotheses concerning advection effects and the thermocline's function as a barrier require some further comment.

The assumed case of negligible advection will only rarely be found in the sea; as a result one can expect deviations from the theoretical equilibrium profile and incorrect predictions of the depth of the thermocline. In our study [1] we proposed an advection diagram along the line from Nice to Calvi, deduced from the local differences between the temperature profile where advection is assumed absent.

The mathematical models proposed assume that the thermocline constitutes an effective barrier. Although the absorption of solar radiation makes itself felt to depths of 100 meters, the effects of surface exchanges with the atmosphere cause and remain confined to, the mixed layer which lies above the thermocline. One thus considers turbulent conduction of heat to become negligible at the level of the thermocline; the experimental examples given confirm this hypothesis. Besides this, J. Gonella [9], from observations of wind, current, and temperature made at the laboratory buoy between 1964 and 1970, has clearly confirmed the barrier effect of the thermocline with regard to momentum. Nonetheless, in a marine environment, characterized

by turbulence this hypothesis only corresponds to a theoretical case and will never exactly describe a real situation. These models for the determination of thermocline depth can thus be no more than estimates, though they have the advantage of being simple and easily checked.

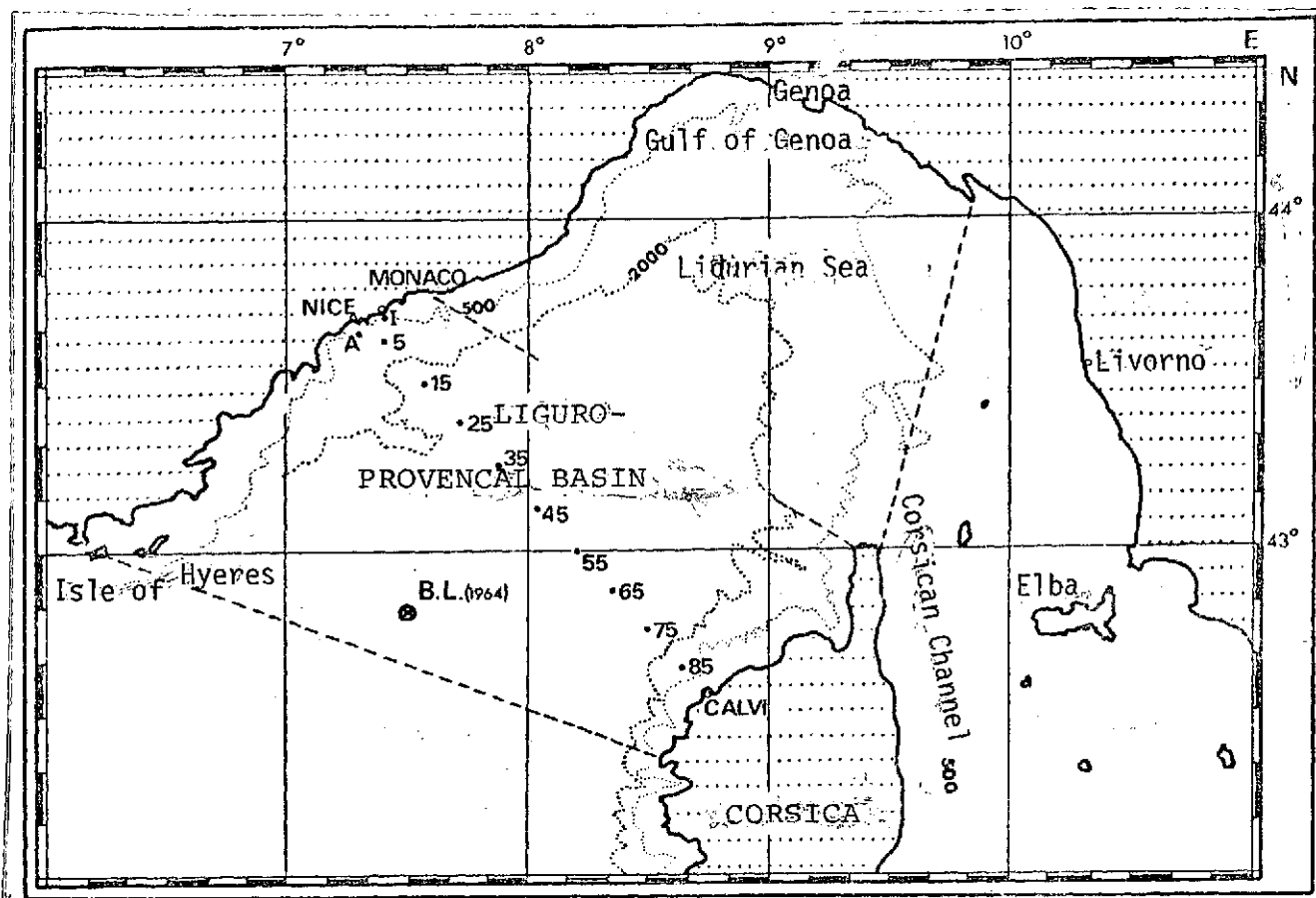


Figure 1

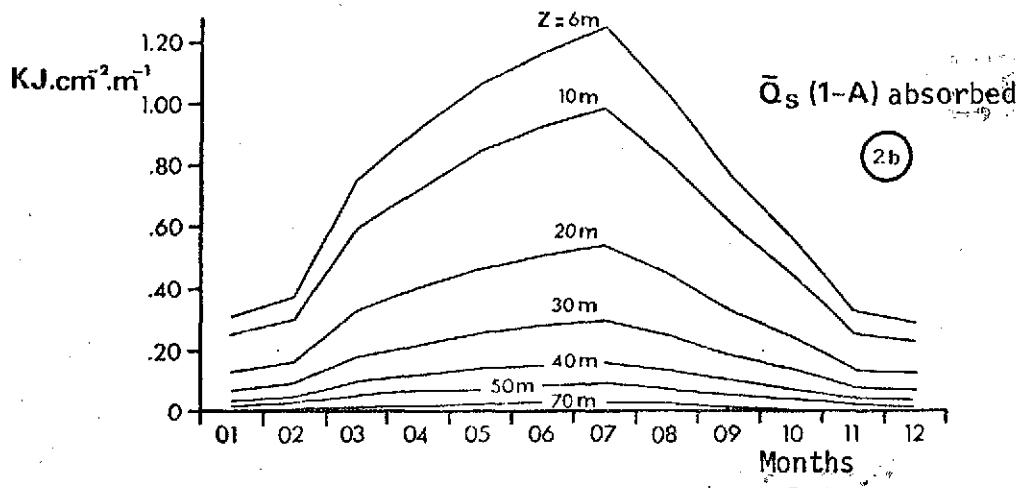
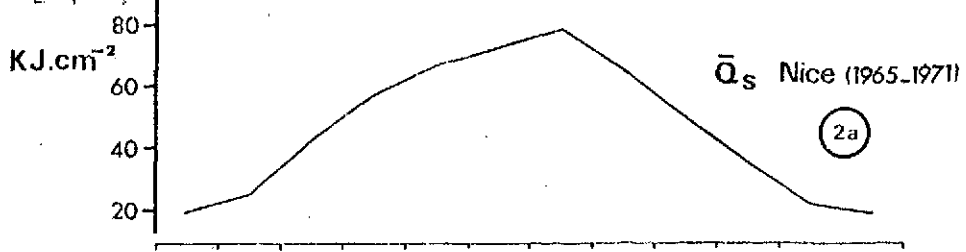


figure 2

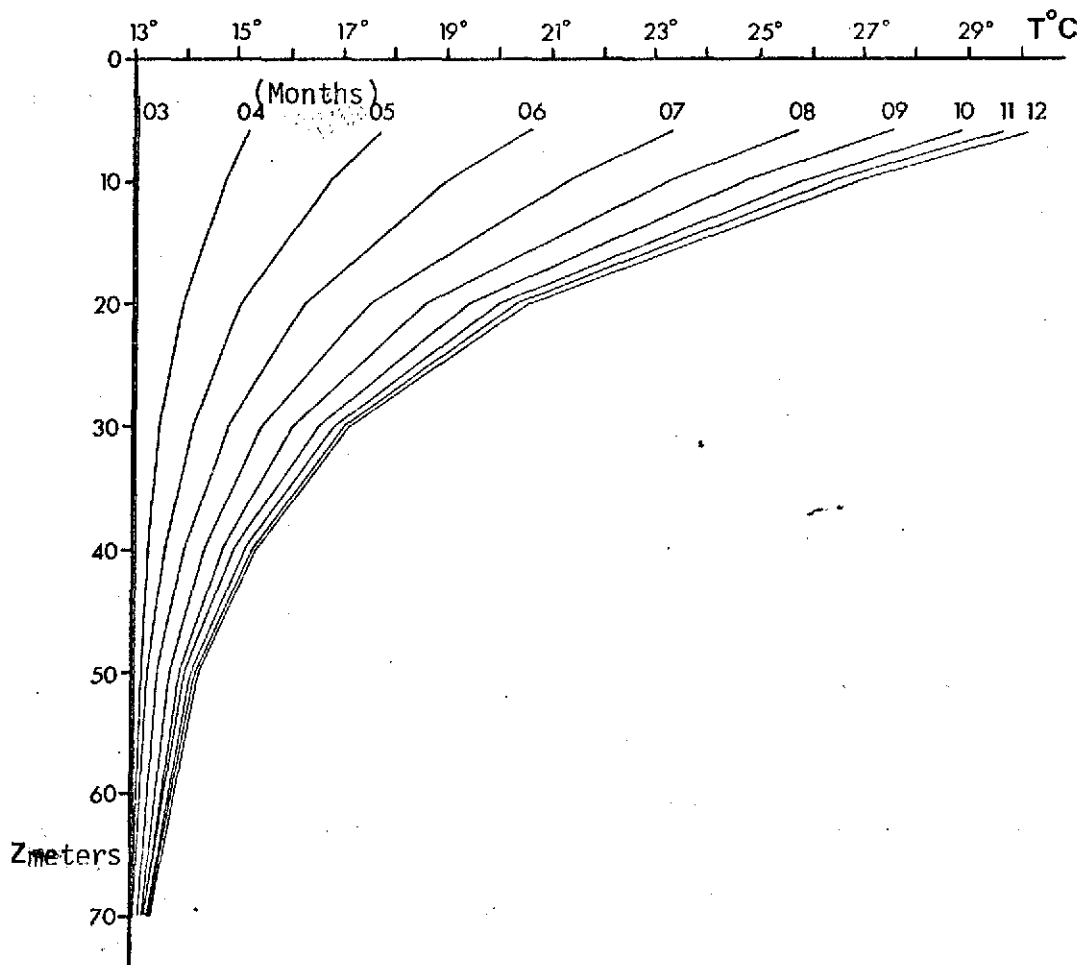
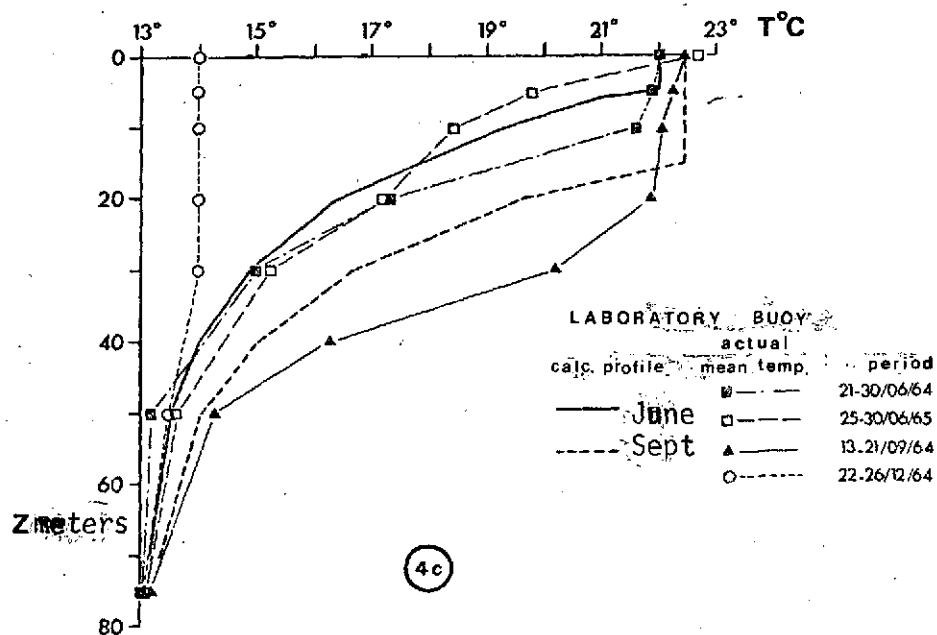
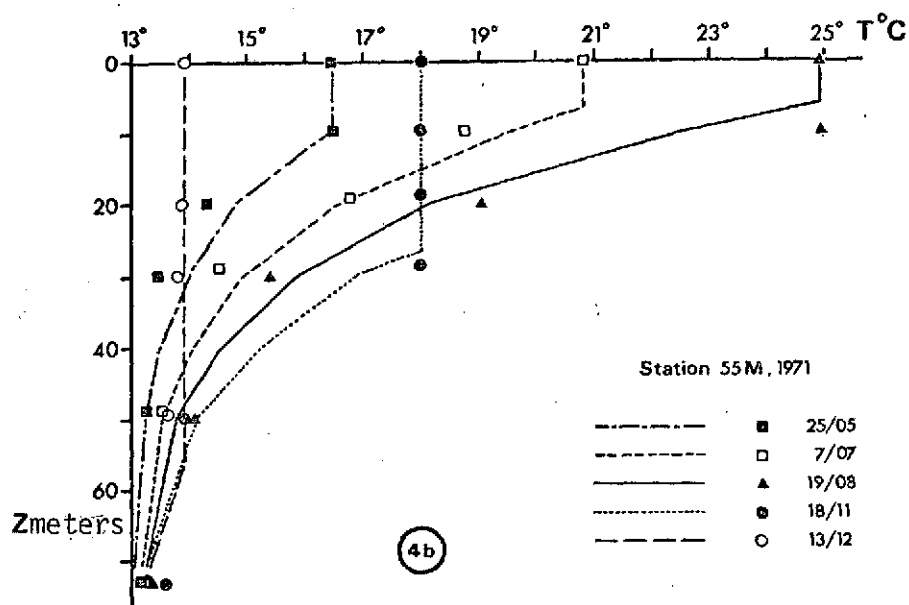
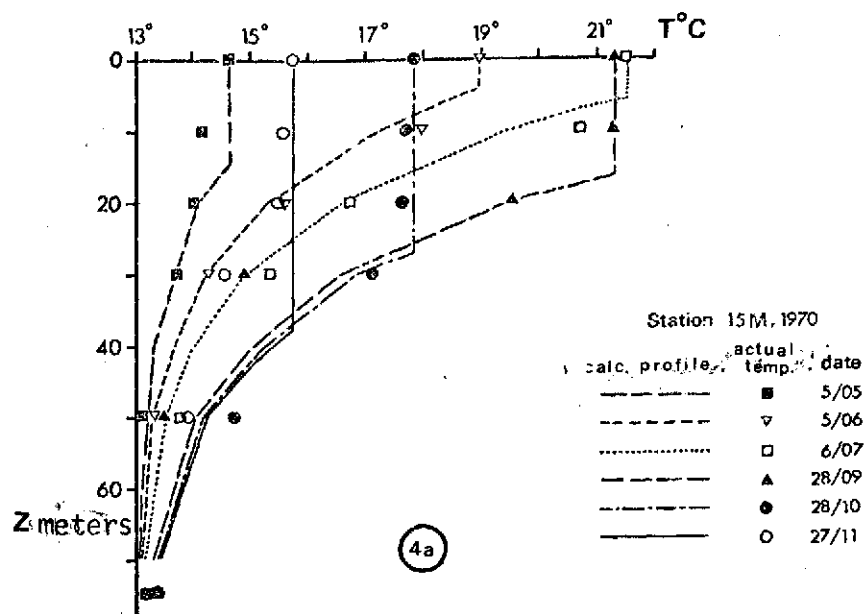


figure 3



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